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Global Volcano Monitoring: What does it mean when volcanoes deform?

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Abstract (97 words)

To look inside a currently active volcano, we use indirect, geophysical methods. One approach is to measure surface deformation, from which we can infer subsurface magmatic or hydrothermal processes. The recent explosion in satellite data means that we can now measure deformation at hundreds of volcanoes without relying on limited ground instrumentation. The number of known deforming volcanoes increased from 44 in 1997 to over 220 as we document here. We review what we have learned from these observations -- the diverse ways that volcanoes can deform, typical rates and durations, and the processes driving that deformation.

Keywords: volcano deformation, eruption, geodesy, InSAR, GPS

Total 3679 words.

1. Introduction

Geophysical methods allow us to study the internal structure of an active volcano and the processes that may ultimately lead to eruption. Geology allows us to see the inner plumbing of old volcanoes that has been exhumed by erosion and the other articles in this issue describe what detailed geochronological and petrological studies of exhumed exposures and eruptive products, such as ash, gas, lava, can tell us about volcanic systems. In contrast, geophysical techniques make measurements near the surface to image the structure beneath a volcano, and measure the stress and strain

associated with magmatic processes (e.g., Pritchard and Gregg, 2016). A common limitation is that multiple sensors must be placed around the volcano in order to make detailed inferences about the subsurface, and the majority of the world's 1400 subaerial volcanoes do not have such networks (e.g., Loughlin et al., 2015). Here we focus on observations of surface deformation ('geodesy') because these measurements can now be made using satellites, and thus globally (e.g., Biggs et al., 2014; Fournier et al., 2010). The number of known deforming volcanoes has quintupled in the last 20 years as traditional ground-based survey methods have been complemented with satellite technology. The ability to observe deformation at a wide range of types of volcanoes, and at different stages of their history, offers a unique perspective on the behavior of magmatic systems throughout the eruption cycle.

Despite the explosion of observations, our understanding of the significance of deformation at volcanoes (or the lack of deformation) is still in its infancy. The challenge for 21st century volcanologists is to link the new observations of surface deformation to volcanic processes, within the framework of other geological and geophysical observations of magmatic systems. We can learn about complex magmatic systems, from the generation of melt to its interaction with the surface, atmosphere, or shallow crust, and improve monitoring and forecasting of volcanic hazards. Here, we will first review the geodetic methods that allow us to study subsurface processes in active volcanoes, and then summarize the broad range of volcanic behavior that has been observed geodetically.

2. Geodetic techniques

In 1997, deformation had been reported at 44 volcanoes (Dvorak and Dzurisin, 1997), by 2010, there were 118 (Fournier et al., 2010) and today there are over 220, which we document in the supplementary table. This rapid increase is not the result of a rise in volcanic activity, but a consequence of our improved observation and reporting, particularly in the developing world. Technological advances have been vital: a fleet of international satellites (10 radar satellites in 2017 and growing, e.g., Pinel et al., 2014) makes routine, global observations, and Global Positioning System (GPS) networks make ground-based geodetic observations routine and affordable. However, the list of deforming volcanoes is still incomplete because some volcanoes have never been studied or only studied incompletely due to inadequate data.

This review focuses on Interferometric Synthetic Aperture Radar (InSAR), the main satellite-based tool that is used to measure surface deformation at volcanoes. Radar (RADio Detection And Ranging) involves the transmission and reception of microwave electromagnetic radiation (roughly 10^8 - 10^{11} Hz or wavelengths of 1 mm to 1 m) (e.g., Pinel et al., 2014). At these wavelengths, radar can see through clouds and most precipitation and do not rely on the sun's illumination, giving them a unique all-weather, day-night capability. Each pixel in a Synthetic Aperture Radar (SAR) image is represented by a complex number, with the amplitude corresponding to the intensity of the returned radar energy and the phase equaling a fraction of the complete wavelength (having a value between 0 and 2π). When the phase from two images are combined to form an interferogram, the phase difference reveals variations in the distance between the ground and the satellite that appear as coloured fringes (Fig. 1). Each fringe corresponds to half the radar wavelength. The rate of ground displacement can be measured by taking the observed change in distance divided by the time interval between SAR images – this interval usually varies between a day and a few weeks and rates of deformation range from mm/yr to several m per day (Fig. 2). Some limitations to the technique are 1) decorrelation, which occurs when characteristics of the ground and its ability to reflect radar waves change rapidly, for example in heavily vegetated or agricultural regions, and 2) atmospheric delays caused by water vapour in the troposphere. Time-series methods combining hundreds or thousands of images are increasingly important in overcoming these limitations; several different approaches and variants to the technique are described in textbooks and review papers (e.g., Dzurisin et al., 2006; Pinel et al., 2014)

The revolution in volcano deformation studies by InSAR comes from the ability to routinely image deformation at nearly all of the world's volcanoes with an accuracy of a few mm to cm. Even so, volcanoes can deform very rapidly (e.g., Fig. 2) and the repeat time between satellite overflights may be too long to capture these temporal changes, particularly as the amount of data available for a given volcano varies widely. In some areas, images are collected during every overpass while in others, data may never have been acquired by certain satellites; furthermore, data from some satellites are available at no cost (such as the European Space Agency's Sentinel mission) while imagery from other satellites can cost thousands. One way to

overcome this limit is to use all available satellites in the international constellation; while the data from the different missions cannot be directly combined, using more satellites increases the frequency of ground observations and thus the ability to detect transient events.

Ground sensors complement InSAR observations, and can be used to evaluate or correct uncertainties. The most commonly applied is the continuous Global Navigation Satellite System (GNSS), which uses signals from the Global Positioning System (GPS) and other navigation satellites to measure three-dimensional changes in co-ordinates. Rather than relying on the transmitted code used by cell phones and other domestic systems, geodetic GNSS receivers use the phase of the transmitted signal to achieve sub-millimetre accuracy (Dzurisin, 2006). Other ground-based systems (tiltmeters, strainmeters, ground-based radar, leveling and triangulation surveys; Dzurisin, 2006) are useful where available, but their use is limited to a handful of volcanoes, and it is not possible to tell how widespread the observed processes are. Ground observations that are collected continuously are especially valuable because they overcome the gaps between satellite InSAR measurements.

3. Diversity of deforming volcanoes

A variety of processes can cause ground deformation at volcanoes – e.g., magma movements, landslides, faults, hydrothermal systems, thermal or thermodynamic volume change from heating, cooling, melting or crystallization (e.g., Dzurisin et al., 2006). But, to some extent, the pattern of ground deformation can be diagnostic of the physical process occurring – a dyke intrusion, cooling lava/pyroclastic flow, or pressurizing “magma chamber” all have distinct patterns of ground displacement viewed with InSAR (Fig. 1). Complexities arise depending on the direction the satellite is looking relative to the ground displacement (e.g., Dzurisin et al., 2006) and when multiple processes occur at nearly the same time (Fig.1f,g). In this section, we describe some commonalities between deforming volcanoes and global patterns in the types of processes that produce ground deformation.

3.1 The classic volcano deformation cycle

According to the classic model of the ‘volcano deformation cycle’, prior to eruption, magma gradually inflates a chamber directly beneath the volcanic edifice, until a

threshold is reached at which point the chamber ruptures, and an eruption rapidly empties and deflates the chamber (e.g., Dzurisin et al., 2006). The inflation phase causes uplift of the ground surface and large numbers of small earthquakes (volcano-tectonic seismicity), while eruption is accompanied by rapid subsidence (Fig. 3a). To the first order, this pattern of co-eruptive subsidence and inter-eruptive uplift has been observed at a number of volcanoes with different characteristic lengthscales and timescales, and used to provide eruption forecasting in a number of cases (e.g., references in Dzurisin et al., 2006; Fournier et al., 2010). However, the recent explosion in monitoring data has demonstrated that the classic model is an oversimplification, and that many volcanoes do not behave in this way. Furthermore, a simple liquid-filled magma is not compatible with geological or petrological observations, as discussed elsewhere in this volume. Nonetheless, the classic model of the volcano deformation cycle remains a useful concept from which to start.

Long-lived eruptions and intrusions provide an alternative opportunity to observe repeated cycles of behavior. For example, the multi-decadal eruptions of Soufriere Hills Volcano (SHV), Montserrat (an andesitic stratovolcano) and Kilauea, Hawaii (a basaltic shield volcano) have served as test-beds for new ideas and new monitoring systems. GPS instruments at SHV observed a simple first-order pattern – when lava is erupting rapidly, the surface subsides, and during periods of no extrusion, it inflates (Wadge et al., 2014). Similar patterns have been observed at volcanoes with multiple, distinct eruptions, such as Fernandina, Galapagos (e.g., Pinel et al., 2014) and Okmok, Alaska (e.g., Lu and Dzurisin, 2014). However, close inspection of the time series, shows that inter-eruptive uplift is interrupted by short reversals that are not associated with any magma output at the surface; at Okmok, these have been attributed to pulses of gas loss or re-absorption (e.g., Caricchi et al., 2014).

3.2 Intrusions

Most magma does not reach the surface – and sometimes the intrusion of dykes and sills, and the growth of plutons are visible in the geodetic as well as geological record. Earthquake swarms at Eyjafallajökull, Iceland in 1994 and 1999 were associated with tens of centimeters of surface uplift, but no eruption (Sigmundsson et al., 2010 and references therein). Similar patterns of seismicity and deformation in 2010 culminated in the eruption that disrupted Europe's air traffic causing huge economic losses.

Magma is supplied in batches, some of which ‘stall’ forming an intrusive complex at the roots of the volcano, while others, as happened in 2010 eruption, reach the surface, triggering large eruptions which can tap multiple reservoirs (Sigmundsson et al., 2010). Intrusions also occur within volcanic edifices; the orientation of these intrusions is controlled by the stress field within the edifice, and if they reach the surface, can determine the geometry of the resulting eruptive vents/fissures (e.g., Bagnardi et al., 2013).

The deformation pattern associated with a dyke intrusion is usually distinctive as it has two lobes of lateral displacement on either side of a zone of subsidence (Fig. 1b); it is clearly different from the circular or elongated bulls-eye pattern of uplift typical of a sill (Fig 1a). Some dykes stall before reaching the surface, while others feed eruptions. Recent examples include the eruption of 1.4 km³ of lava at Holuhraun, Iceland which was fed by a dyke originating at Bardarbunga caldera 45 km away (Sigmundsson et al., 2014) and an 80-km long dyke in Afar, Ethiopia, which intersected two small silicic centres triggering minor eruptions and a number of shorter dykes that failed to reach the surface (Pinel et al., 2014 and references therein).

Continental growth also occurs by the formation of plutons and associated ductile deformation, which are major geological features, but are harder to observe geodetically, as they are thought to grow slowly over millions of years. Possible candidates for these processes that could be on the order of mm/yr spanning 10-100 km include Uturuncu in the Central Andes, coincident with the mid-crustal Altiplano-Puna magma or mush body (e.g., Pritchard and Gregg, 2016) and uplift lasting more than 100 years at Socorro, New Mexico (Pearse and Fialko, 2010).

3.3 Volcanic Conduits

Volcanic conduits are thought to be extremely narrow (<10m radius) and are therefore difficult to observe geophysically. Nonetheless, clues to their behavior are given by sub-daily deformation cycles that are typically observed in association with Strombolian or Vulcanian eruptions. Tiltmeters installed on the crater rim at SHV in 1996-1997 detected deformation patterns correlated with seismicity and explosions (Voight et al., 1999). The debate continues as to whether these observations should be

interpreted as pressurization of shallow reservoirs, shear along the conduit wall, or the passage of volatiles (e.g., Nishimura, 2009).

Lava lakes are essentially magma-filled conduits open at the surface, and in a simple conceptual model, changes in the reservoir pressure would result in changes in lake level rather than surface deformation. Many of the world's lava lakes happen to be in remote or inaccessible locations, but the best studied is undoubtedly Halemaumau, Hawaii, where Patrick et al (2015) show that both deformation and lake level change occur simultaneously -- evidence for an open conduit linking the lava lake to the magma reservoir.

3.4 Restless Calderas.

Caldera systems have long repose periods between very large eruptions, but do not remain quiescent – many calderas have frequent, small eruptions, known as resurgent volcanism. In the case of Santorini, Greece, the volume of erupted lava is directly proportional to the time since the previous eruption, evidence that magma supply from deep is continuous (Parks et al., 2012). However, in 2010, a short period of rapid uplift occurred with an equivalent volume to that anticipated for the next eruption, indicating that the magma supply to the shallow reservoir is pulsed rather than continuous (Parks et al., 2012).

Some caldera systems have been known to be deforming for decades without erupting. At Campi Flegrei, Italy, gradual subsidence over centuries has caused the Roman market at Pozzuoli to become submerged. The subsidence is occasionally interrupted by pulses of uplift but it remains unclear whether the cause is magmatic or hydrothermal (e.g., Chiodini et al., 2010). At Yellowstone, deformation has a spatially and temporally variable pattern and is attributed to hydrothermal fluids moving between reservoirs as well as magma intrusion (e.g., Chang et al., 2007).

More recently, satellite observations have shown that deformation is occurring at many caldera systems that are not erupting, have no record of historical volcanism and no real-time monitoring. This includes many volcanoes along the densely populated East African Rift (e.g., Biggs et al., 2014 and references therein) and raises the question of how to interpret caldera deformation in terms of volcanic hazards. The

fastest, Laguna del Maule, Chile, has been uplifting at a rate of 280 mm/yr since 2007 (e.g., Fournier et al., 2010) but the implications for the shallow magma body underneath remain unclear.

Is the behavior of restless calderas related to external influences? The passage of seismic waves associated with large earthquakes has been shown to influence hydrothermal systems, probably as a result of interactions with gas bubbles. For example, both the 2010 Maule, Chile and 2011 Tohoku, Japan earthquakes caused subsidence at several nearby volcanoes, but the mechanism could be either changes in the hydrothermal system or stress changes acting on the weaker rheological properties of a pluton (e.g. Pritchard et al., 2013).

3.5 During and after the eruption

While co-eruptive subsidence associated with subsurface magma withdrawal is a common feature of eruptions, in many cases it is masked by local surface changes, including both the destruction and creation of topography. Explosive eruptions typically form new craters and vents, or alter the morphology of existing structures, and may trigger large collapses. Eruptive products including lava flows, pyroclastic flows and lahars fill low topography, while a new dome may grow near the vent (e.g., Pinel et al., 2014). Ash fall, while rarely thick enough to dramatically alter topography, can alter the appearance of the ground surface making it incoherent to satellite radar and reducing the efficiency of ground-based monitoring that relies on solar panels. After the eruption, the new topography is typically oversteep, and processes such as landsliding (e.g., Ebmeier et al., 2014) and gravitational spreading (e.g., Schaefer et al., 2015) act to restabilise the landscape. Subsidence of cooling and compacting lava flows can last for decades, as seen at Paracutín, which erupted in 1952 and still subsides today (e.g., Fournier et al., 2010).

Magmatic systems can respond to eruption in multiple ways – in the classic model of the volcano deformation cycle, the magma chamber begins to refill within days of the eruption (e.g., Okmok; Lu and Dzurisin, 2014) but elsewhere, the volcano continues to subside (e.g., Kiska; Lu and Dzurisin, 2014). For decades after an intrusion, the subsurface magma body will continue to cool, crystallize and degas: at Medicine Lake Volcano in the Cascade Range, combining modern geodetic observations with

leveling surveys from the 1950s demonstrate that subsidence has continued at ~10 mm/yr for at least 65 years (Parker et al., 2014).

3.6 Limits of detectability

Several eruptions appear to have taken place at volcanoes with no known deformation (e.g., Fournier et al., 2010; Lu and Dzurisin, 2014), and even when co-eruptive subsidence is observed, the volume is usually less than the dense rock equivalent of the erupted products. The classic model of the volcano deformation cycle assumes a constant volume flux is supplied to the magma chamber, producing a linear rate of uplift between eruptions. Altering the boundary conditions such that magma is supplied from a deeper chamber at constant pressure modifies this model such that eruption is followed by exponentially decaying uplift. This may explain observations of eruptions that occur with negligible uplift prior to eruption and very rapid uplift following eruption, such as Westdahl, Alaska (e.g., Lu and Dzurisin, 2014).

The simplest explanation for a lack of co-eruptive subsidence is that deformation is occurring at a magnitude or resolution beneath our current ability to observe: too fast or too slow, too shallow or too deep, or obscured by atmospheric effects. Already, new satellites with higher resolution satellites and faster repeats (e.g. TerraSAR-X and CosmoSkyMed) have been used to observe previously-undetectable processes (e.g. Salzer et al., 2014) and long-term monitoring missions, such as the Sentinel programme, have the potential to revolutionise detection capabilities. An alternative explanation is that we ought to consider mass, rather than volume balance, as density is far from constant in a three-phase magmatic system undergoing changes in temperature and pressure and gas bubbles form very compressible magmas (e.g.,Caricchi et al., 2014).

4. Global Synthesis

The rapid increase in the number of geodetically-studied volcanoes means that it is now feasible to treat the observations statistically. Biggs et al (2014) showed that there is a strong link between deformation and eruption for over 500 systematically studied volcanoes, and also between non-deformation and non-eruption. Further, this relationship varies with volcanic parameters: shield volcanoes have a strong link between deformation and eruption and come closest to the classic volcano

deformation cycle, while calderas frequently deform without erupting, suggesting that large volumes of magma are stored in the upper crust, and the deformation is associated caused by gas or hydrothermal fluids. Stratovolcanoes are the most likely to erupt without observable deformation, perhaps because the mass changes are accommodated in a way not detectable by current observation systems.

How do we know when a given deformation event may lead to eruption? One way to answer this question is to compare the duration of deformation events to their magnitude and whether or not they led to eruption (Fig. 2). As expected, deformation events that have large magnitudes usually can only last for a short amount of time and typically lead to eruption (e.g., Fournier et al., 2010), but there doesn't seem to be a simple threshold at which deformation duration and/or magnitude should cause concern. On the other hand, there are many volcanoes that can deform at rates of 1-1000 mm/yr without causing eruption in the short term at least, and so these types of deformation events are not always hazardous. This figure and supplementary table are incomplete in several respects – some volcanoes have never been studied, at others the temporal sampling is inadequate to resolve pulses of rapid deformation, the deforming area may have been too small to resolve or occurred within a data gap, and in some cases deformation was ongoing at the start or end of the available observations, so the specified duration is a minimum estimate. There is a cluster of deformation events with a 1 day duration that likely correspond to even shorter time periods but could not be constrained because observations were not sufficiently frequent in time.

Although deformation has been reported at over 200 volcanoes, there remain several eruption styles that have been observed rarely, or not at all. The only geodetically-observed rhyolite eruption occurred at Chaitén, Chile in 2008 with only a few hours of pre-eruptive warning (e.g., Wicks et al., 2011), but it is impossible to state whether the pattern observed is representative. The latest monogenetic eruption occurred at Paracutín in 1943-1953, before routine satellite observations. Perhaps most alarming is the lack of observation of the very largest eruptions. The latest magnitude 7 eruption (a volume in dense rock equivalent of 10^{11} - 10^{12} km³) occurred at Tambora, Philippines in 1815, and even recent magnitude 5-6 eruptions, with erupted volumes of 10^9 - 10^{11} km³, such as Mt. St. Helens, and Mt. Pinatubo, have limited geodetic

observations. The only available example is the relatively small eruption of Campi Flegrei in 1538, where historical records suggest that several metres of deformation occurred in the years before the eruption (Guidoboni and Ciuccarelli, 2011). Calderas have the potential for extremely large eruptions, but could we distinguish between the semi-continuous unrest seen at so many calderas and the precursors to a major eruption?

5. Summary and Perspectives

Satellite techniques are rapidly improving our ability to monitor volcanoes on a global basis, and have provided insight into the spatial and temporal changes in the subsurface stress fields around volcanoes and intrusions. The classic model of the volcano deformation cycle - co-eruptive deflation and inter-eruptive inflation - is seen at many volcanoes, but the rapid increase in geodetic monitoring has demonstrated that such a simple model is not always applicable. Deformation can be attributed to the movement and phase transitions of magma, volatiles and hydrothermal fluids and intrusions of dyke and sills and the growth of plutons, each of which have their own characteristic geodetic signatures. Surface processes during and after eruptions such as the formation of volcanic flows and domes, and edifice instability may obscure subsurface processes. Alongside developments in observational ability, modeling capabilities have evolved from analytic solutions for point sources and other simplified geometries, to finite element models incorporating more complex rheological and structural information, to physics-based models capable of integrating geophysical monitoring with observations of degassing, petrology and so on (Masterlark, 2006; Anderson and Segall, 2011).

Satellite imagery has proved a remarkable reconnaissance tool to discover new phenomena, and as new satellites and constellations of satellites are launched in the coming years (e.g., Pinel et al., 2014), even more new discoveries will be made. Yet, the biggest question remains: which of these deforming volcanoes are a threat, and under what circumstances does deformation lead to eruption?

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Figure 1: Some examples of the diversity of volcano deformation patterns seen with InSAR. Images are ‘wrapped’, and each coloured ‘fringe’ can be thought of as a contour line – the total displacement can be calculated by counting the number of fringes. A) Deformation at Lazufre on the Chile/Argentina border (Fournier et al., 2010) B) Dyke intrusion in Afar, showing characteristic two-lobed pattern (Hamling et al., 2009). C) Flow subsidence at Tolbachik Volcano, Russia; deformation is irregular and restricted to the extent of the deposit (Fournier et al., 2010) D) Shallow landslides at Arenal Volcano, Costa Rica (Ebmeier et al., 2014). E) Subsidence associated with cooling and crystallization of an intrusion at Medicine Lake Volcano, Cascades. The same pattern is seen in leveling surveys extending back 60 years (Parker et al., 2014). F) Broad surface uplift (70 km diameter) surrounded by a ring of subsidence (150 km diameter) centered on Volcán Uturunu, Bolivia (e.g., Pritchard and Gregg, 2016). G) Multiple horizontally-separated deformation sources in Hawaii, showing simultaneous subsidence of a magma reservoir under Kilauea caldera and dyke intrusion into the East Rift Zone. M.Poland, personal communication.

Figure 2: Rates of volcano deformation as a function of the duration of the deformation from the global compilation of 485 events from 221 unique volcanoes in supplemental material. Deformation magnitudes indicate maximum displacements (horizontal or vertical for GPS, along the line of sight in the case of InSAR), while the duration indicates the time (or best estimate of the time) that event took to unfold. Points are colored “no eruption” or “eruption” to indicate where the deformation occurs before, during, or immediately after an eruption or is not obviously related to eruption.

Figure 3: Temporal patterns of volcano deformation (schematic); stars represent eruptions. A) Classic model of the eruption cycle where emptying of a magma reservoir during eruption causes subsidence, and refilling of the reservoir at a constant rate between eruptions causes uplift at a constant rate. B) Modified eruption cycle, such that the rate of refilling between eruptions decays exponentially as magma flows along a pressure gradient from a deep reservoir. C) Magma rises quickly and erupts immediately, such that deformation is rapid, recoverable and is likely to be undetected. D) Pulsed magma supply, batches of magma are intruded causing uplift, until a threshold is reached and an eruption is triggered. E) Continuous unrest without eruption, caused by phase changes and mixing within shallow magma storage, and possibly the overlying hydrothermal system. F) Constant rate uplift or subsidence which may continue for several decades as a consequence of the growth or cooling of deep magma bodies.